THE ROLE OF BASALTIC VOLCANISM IN LUNAR EVOLUTION: TESTING MODELS OF PETROGENESIS. **James W. Head, Department of Geological Sciences, Brown University, Providence, RI 02912 USA (james_head@brown.edu)**

Introduction: The record of volcanism on the Moon is a crucial component to understanding its geologic, geodynamic and thermal evolution. Using Galileo and Clementine data we have focused on the redefinition of mare units and their assessment in the nearside maria [e.g., 1], the definition and characterization of local and regional dark mantle units [2], the definition, mapping, and comparison of mare units associated with regional and local volcanic complexes [3]. Early on, basic models for the formation of mare basalts involved internal radiogenic heat sources and sequential partial melting of a layered mantle. Subsequent petrological evidence suggested a wide range of source depths for mare basalts. The documentation of increasing complexities in the ages and petrogenesis of mare basalts led to several major new ideas for their generation (see summary in [4]). Here we outline the major models that have been proposed for the origin of lunar mare basalts in order to begin to test these models with the accumulated observations.

Model 1: The nearside/farside asymmetry in the distribution of mare basalts is due to crustal thickness differences: One of the more fundamental characteristics of the Moon is the nearside-farside asymmetry in the distribution of mare basalts. Early models accounted for this by calling on differences in NS/FS crustal thickness [5-6]: for globally heterogeneous mare basalt source regions at depth, dikes rising to a constant level were much more likely to erupt onto the surface in the thinner nearside crust than dikes rising in a thicker farside crust. The general paucity of farside mare basalts, and the concentration of most farside basalts in the relatively deep South Pole-Aitken basin seemed consistent with this hypothesis. Global altimetry obtained by the Clementine mission [7-8] revealed, however, that the depth of the South Pole-Aitken basin was comparable to mare basalt elevations in basins on the nearside, and yet the SPA basin was not as extensively flooded by mare basalts as the nearside basins. These observations cast doubt on the assumption of global mare basalt source symmetry and the role of crustal thickness in explaining NS/FS mare basalt asymmetry, and pointed out the necessity of adding additional factors to models of magma ascent and eruption [9]. This led to the development of several alternative models for the emplacement of basalts in the South Pole-Aitken basin and for the Moon as a whole raising the questions: were there fundamental differences in the nearside-farside source regions established early in lunar history; did the Procellarum KREEP Terrain (PKT) play a fundamental role in mare basalt generation; was there a basic difference in the nearside/farside thermal gradient that influenced both basin relaxation and mare basalt generation, ascent and eruption; and could the formation of a basin the size of SPA have induced sufficient convection to have stripped away a subsurface KREEP layer, and thus to have inhibited the formation of mare basalts below the basin?

Model 2: Impact basin pressure-release melting and associated secondary convection explain the observed distribution of mare basalts: Pressure-release melting is known to be an important mechanism for basalt generation, but the lunar pressure gradient, combined with the composition of the crust and the apparent depth of origin of mare basalts, led its initial disfavor. Recently [10] have reassessed the magmatic effects of large basin formation, and introduced a two-stage model for melt creation beneath lunar basins triggered by basin formation itself. In the initial stage, crater excavation depressurizes and uplifts underlying mantle material so that it melts in-situ instantaneously, forming large quantities of melt below the basin (in addition to impact melt in the cavity). This model thus predicts huge quantities of in situ pressure-release melt (98-100% of the melt created by both mechanisms) produced instantaneously and available to be extruded into the impact basin as lunar mare basalts. In the second stage, the cratered lithosphere rises isostatically, warping lithosphere-asthenosphere isotherms upward and inducing convection, at which time adiabatic melting can occur. This second stage produces only ~1-2% of the total melt but can last for a longer period of time, up to ~350 Ma. In the model [10], mafic mantle melts can be generated from depths of 150-560 km, depending on mantle potential temperature. Assuming that 10% of the melt generated erupts, [10] find that the volumes of magma reported for basins are similar to their predictions (their Fig. 5). They [10] also model the origin and emplacement of high-alumina, high-TiO_2, KREEP-rich and picritic magmas, and predict an order of eruption, with the most primitive, lowest titanium magmas last. In a more recent treatment of impact-induced convection, [11] model basins of different sizes (e.g., ranging from Orientale, through Imbrium, to South Pole-Aitken), make different predictions about the record of mare basalt emplacement for each, and these can be readily tested against the geological record of mare basalt volcanism that we have compiled. Key tests involve the timing, duration, volumes, styles and the comparison of the record in different sized basins.

Model 3: An enhanced KREEP layer in the Western Nearside Procellarum KREEP Terrain (PKT) explains the generation, distribution, and mode of emplacement of mare basalts: More than 60% of the mare basalts by area occur within the boundaries of the Procellarum KREEP Terrain (PKT) which makes up only ~16% of the surface of the Moon. This observation was one of the major factors that lead [12] to propose that there was a cause and effect in that the enhancement of KREEP heat-producing elements significantly influenced the thermal evolution of the region, causing the underlying mantle to partially melt over much of lunar history to generate the observed basaltic volcanic sequence. The thermal model of [12] predicts that partial melting begins immediately after the model is started at 4.5 Ga and continues to a lesser degree to the present, melting initiates immediately beneath the KREEP basalt layer and becomes deeper with time, with the maximum depth of melting being ~600 km, and the KREEP layer is kept above its liquidus for most of lunar history. Does this hypothesis account for the origin of mare basalts in terms of the timing, duration, areal distribution, volumes, and changes in depth with time? Others [13] pointed out several difficulties with this model in terms of petrogenetic evolution, geophysical evidence against the long-term duration of a near-liquid KREEP layer, and the fact that such a layer might form an impenetrable barrier to the eruption of mare basalts. To date, no one has undertaken an analysis of the nature of volcanism in the PKT to address the predictions of the model [12] and the caveats [13].

Model 4: The large-scale overturn of an initial unstable stratification causes the generation of mare basalts: In this model [14] , crystallization of the lunar magma ocean (LMO) forms a chemically stratified lunar interior with an anorthositic crust separated from the primitive lunar interior by magma ocean cumulates. Dense, ilmenite-rich cumulates with high concentrations of incompatible radioactive elements are the last magma ocean cumulates to form, and underlying olivine-orthpyroxene cumulates are also stratified with later crystallized, denser, more Fe-rich compositions at the top. These layers are gravitationally unstable. Rayleigh-Taylor instabilities cause the dense cumulates to sink toward the center of the Moon and to form a dense core. Subsequently, the ilmenite-rich cumulate core undergoes radioactive heating and this heats the overlying mantle, causing melting. The source region for high-TiO_2 basalts is thus envisioned to be a mixed zone above the core-mantle boundary containing variable amounts of ilmenite and KREEP, and involves deep, high pressure melting, delayed for a period of time subsequent to LMO formation and overturn. Thermal plumes rise into chemically stratified surroundings of the mantle (chemically less dense but colder) above the core and cause mixing and homogenization. The resulting lower thermal boundary layer may be partially to wholly molten depending on mineralogy and the range of input parameters. Melting at the top of the mixed layer to produce mare basalt magmas must occur at low enough pressure for melt buoyancy and at high enough pressure to satisfy the depth indicated by phase equilibria. The onset time of mare volcanism is constrained by bulk core radioactivity, and TiO_2-rich mare basalt liquids must be positively buoyant enough to form dikes rather than sink. Does the distribution in time and space and the mode of emplacement derived from our synthesis stratigraphy and vent characteristics support or refute the overturn model?

We use these predictions and the documented stratigraphy to test these models.

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